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DIURNAL WIND VARIATIONS, SURFACE TO 30 KILOMETERS

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ABSTRACT

Annual mean 12-hr wind differences are computed from monthly mean wind statistics at 27 levels between the surface and 10 mb for 105 radiosonde stations that have extended periods of record. The resulting wind difference vectors for 0000–1200 GMT and 0300–1500 GMT are plotted on constant pressure charts. At polar latitudes in both hemispheres, a simple pattern is observed, with flow directed across the Pole from the daytime hemisphere to the nighttime hemisphere at all levels. At low and middle latitudes, the wind difference patterns are strongly related to topography, even at stratospheric levels. Land-sea contrasts and terrain slope appear to be the controlling influences.

Analysis of hodographs at individual stations shows that the topographically induced tidal fluctuations propagate downward at all levels above 1 km. This is indicative of an upward flux of tidal energy from the planetary boundary layer.

1. INTRODUCTION

The subject of the diurnal tide in the upper stratosphere and mesosphere has been actively pursued during recent years. Rocket data have revealed large-amplitude fluctuations above 30 km that can be described in terms of coherent large-scale patterns. Recent developments in tidal theory by Lindzen (1967) and others have explained most of the major features observed at these levels as a response to absorption of solar energy by water vapor and ozone.

Below 30 km, progress in this area has been much more limited despite the relatively large amount of conventional radiosonde data available for studying the tide. The small amplitudes, and the diversity of results, which have thus far been obtained, have tended to discourage further observational studies. Large amounts of data are required to obtain statistically significant results for a particular station, and often it is difficult to relate the results for individual stations to the global pattern.

The first published attempt to investigate the diurnal variation of winds in the portion of the stratosphere below 30 km was made by Johnson (1955). By combining series of 6-hr observations made at different times, he computed the diurnal and semidiurnal wind variations in the lower stratosphere over the British Isles. Tests for significance of the results showed the variations to be real and not random fluctuations. The diurnal wind variation was found to be the same amplitude as the semidiurnal wind variation. Studies by Harris (1959) and Harris, Finger, and Teweles (1962) contrasted tidal fluctuations in wind and temperature at Lajes Field, Azores, with those at Washington, D.C. Although the stations lie at approximately the same latitude, the wind fluctuations at Washington were considerably larger, with peak amplitudes at 1 and 5 km. In an expanded study, combining data from eight stations spanning a large range of longitudes, Finger, Harris, and Teweles (1965) computed the phase of the diurnal tide. Their combined results showed a maximum poleward wind component at about local noon and a maximum westerly component about 6 hr later at all levels.

In contrast to this relatively simple conception of the tidal wind pattern, Hering and Borden's (1962) study of summer wind data over the United States revealed a complex behavior. Amplitude and phase of the diurnal cycle were found to vary greatly with height, with stations in the Great Plains showing large peak amplitudes near 1, 5, and 12 km. Wind vectors at individual stations exhibited a uniform clockwise rotation with time at each level, and there appeared to be horizontal consistency between the wind vectors at neighboring stations. Maps of the instantaneous wind vectors at the 1-, 5-, and 12-km levels showed a suggestion of large-scale patterns that appeared to be related to the topography.

Rasmusson (1967), in his study of water vapor transport over the North American Continent, had occasion to comment upon diurnal differences in the summer wind field. He noted that the large amplitude patterns found by Hering and Borden (1962) over the Great Plains Region extended southward to the Yucatan Peninsula and eastward through much of the southeastern United States and the Caribbean, thus affecting all the regions bordering on the western extension of the Atlantic subtropical High. Stations on the Pacific coast showed evidence of the influence of another large-scale diurnal wind system. Lettau (1967) has shown evidence of another strong diurnal circulation pattern in the lower troposphere along the western slopes of the Andes.

The purpose of the present study is to extend the investigation of the diurnal wind variations below 30 km to the limits of the present available data coverage.

2. DATA

The data used in this study were supplied by the U.S. Navy Weather Research Facility, processed in the form of mean monthly winds for 32 pressure levels between the

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TABLE 1.—*List of stations (Northern Hemisphere)*

Station	Latitude (° N.)	Longitude (° W.)	WMO No.
Alert	82.5	62.3	74082
Anchorage	61.2	149.9	70273
Annette	55.0	131.6	70398
Baker Lake	64.3	96.0	72926
Balboa	9.0	79.6	78806
Barrow	71.3	156.8	70026
Barter Is	70.1	143.7	70086
Bermuda	32.4	64.7	78016
Bethel	60.8	161.7	70219
Bismarck	46.8	100.7	72764
Bogotá	4.7	74.2	80222
Buffalo	42.9	78.7	72528
Caribou	46.9	68.0	72712
Charleston	32.9	80.0	72208
Chateauroux	46.9	1.7 E.	07354
Churchill	58.7	94.1	72913
Columbia, Mo	39.0	92.4	72445
Coppermine	67.8	115.1	72938
Edmonton	53.6	113.5	72879
El Paso	31.8	106.4	72270
Ely	39.3	114.9	72486
Eniwetok	12.3	162.3 E.	91250
Eureka, NWT	80.2	86.2	72917
Fairbanks	64.8	147.9	70261
Frobisher	63.7	68.5	72909
Ft. Chimo	58.1	68.4	72906
Ft. Smith	60.0	112.0	72934
Goose Bay	53.3	60.4	72816
Grand Bahama	26.6	78.3	78063
Grand Cayman	19.3	81.4	78383
Grand Junction	39.1	108.5	72476
Great Falls	47.5	111.3	72775
Guam	15.0	144.8 E.	91217
Guantanamo	19.9	75.1	78367
Hilo	19.7	155.1	91285
Isachsen	78.8	103.5	74074
Iwo Jima	24.8	141.3 E.	91115
Johnston Is	17.3	169.5	91275
Keflavik	64.0	22.6	04018
Kenitra	34.3	6.6	60119
Kingston	17.9	76.8	78397
Kwajalein	8.7	167.7 E.	91366
Lajes	38.7	27.1	08509
Lihue	21.9	159.3	91165
Luzon	15.2	120.6 E.	98327
Maniwaki	46.4	76.0	72722
Medford	42.4	122.9	72597
Merida	21.0	89.5	76644
Midway	28.2	177.3	91066
Moosonee	51.3	80.6	72836
Mosulpo	33.2	126.2 E.	47187
Mould Bay	76.2	119.3	74072
Nantucket	41.2	70.1	72506
Nashville	36.1	86.7	72327
Nome	64.5	165.4	70200
Norman Wells	65.3	126.8	74043
Oakland	37.7	122.2	72493
Ocean Sta. B	56.5	51.0	02B
Ocean Sta. C	52.8	35.5	03C
Ocean Sta. D	44.0	41.0	04D
Ocean Sta. E	35.0	48.0	05E
Ocean Sta. N	30.0	140.0	24N
Ocean Sta. P	50.0	145.0	17P
Ocean Sta. V	34.0	164.0 E.	25V
Okinawa	26.3	127.7 E.	47931
Oklahoma City	35.4	97.6	72353
Omaha	41.4	96.0	72553
Osan	37.1	127.0 E.	47122
Plessman	12.2	69.0	78988
Port Harrison	58.4	78.1	72907
Prince George	53.9	122.7	72896
Rapid City	44.1	103.1	72662
Resolute Bay	74.8	95.0	72924
Sable Is	43.9	60.0	72600
Sach's Harbor	71.9	124.7	74051
San Antonio	29.5	98.5	72253
San Diego	32.7	117.2	72290
San Juan	18.5	66.1	78526
Santo Domingo	18.5	69.4	78486
Sault Ste. Marie	46.5	84.4	72734
Shreveport	32.5	93.8	72248
St. Martins	18.0	63.1	78866
St. Paul Is	57.1	170.2	70308
Swan Is	17.4	83.9	78501
Tatoosh	48.4	124.7	72798
The Pas	54.0	101.1	72867
Thule	76.5	68.8	04202
Trinidad	10.7	61.4	78967
Tripoli	32.9	13.3 E.	62011
Trout Lake	53.8	89.9	72848
Wake	19.3	166.6 E.	91245
Washington	38.8	77.0	72405
Yakutat	59.5	139.7	70361
Zaragoza	41.7	1.1	08159

TABLE 2.—*List of stations (Southern Hemisphere)*

Station	Latitude (° S.)	Longitude (° W.)	WMO No.
Amundsen-Scott	90.0	----------	89009
Antofagasta	22.5	70.4	85442
Byrd Station	80.1	119.5	89125
Ellsworth	77.7	41.1	89043
Hallett	72.3	170.2 E.	89671
Lima	12.0	77.1	84628
Little America	78.2	162.2	89162
McMurdo	77.8	166.7 E.	89664
Puerto Montt	41.5	72.9	85801
Quintero	32.8	71.5	85543
Wilkes	66.2	110.5 E.	89611

surface and 10 mb. The stations listed in tables 1 and 2 had twice-daily wind data available for part or all of the period 1950–64. Prior to June 1, 1957, upper air observations were taken at 0300 and 1500 GMT; since that time the observation times have been 0000 and 1200 GMT.

Monthly mean 12-hr wind differences were computed for each month. In order to reduce the randomness due to missing data, monthly mean wind differences composed of fewer than a specified number of observations at either observation time were discarded. At levels up to 100 mb, the lower limit was set equal to 25 observations per month. It was felt that at and above 100 mb, where the synoptic situation changes slowly from day to day, there was justification for retaining the mean monthly wind differences based on 18 or more observations at both times.

Mean annual wind differences were then formed for each station by averaging all the mean monthly differences at each level. In the averaging process the 0300–1500 GMT and the 0000–1200 GMT data were treated separately. A similar averaging was performed to compute mean seasonal wind differences. The 12-hr wind differences thus computed contain contributions from all the odd harmonics of the diurnal cycle, but these should generally be small in comparison to the diurnal tide itself.

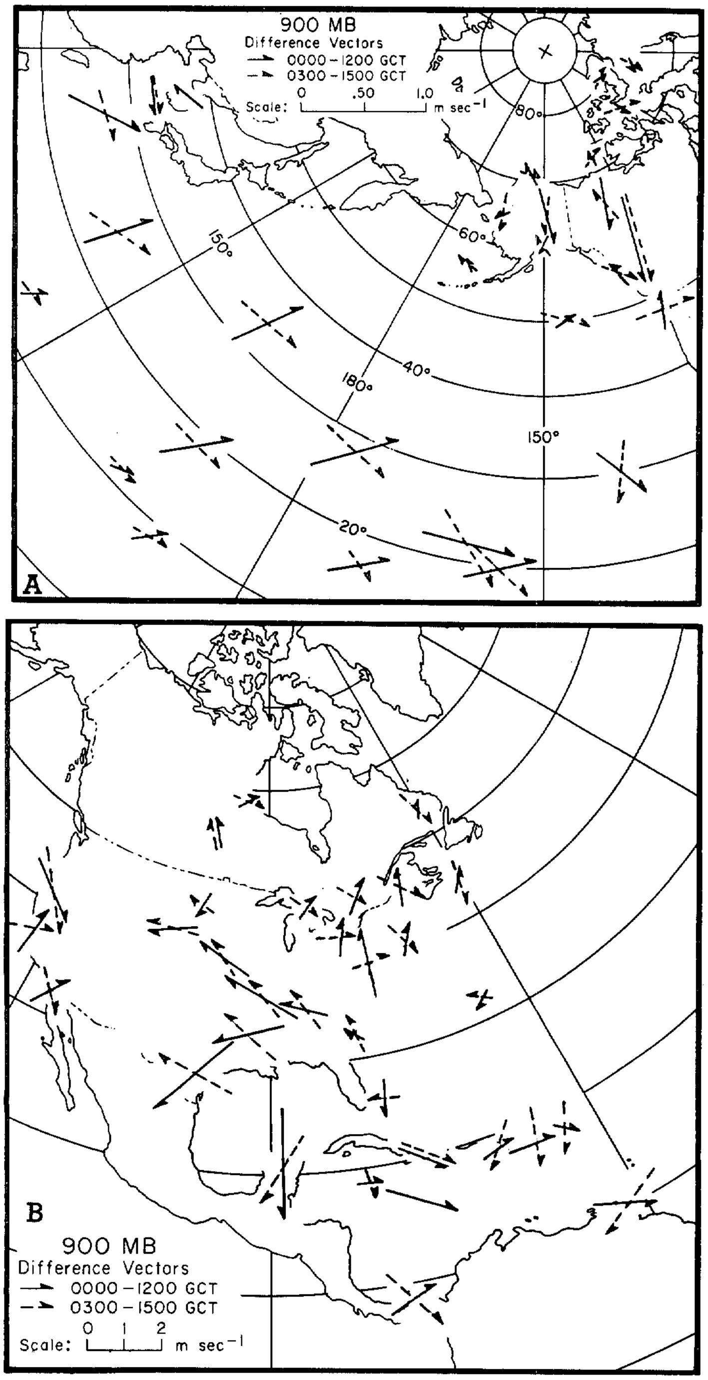

FIGURE 1.—Annual average 0000–1200 GMT (solid) and 0300–1500 GMT (dashed) wind differences at 900 mb, plotted in vector form.

Hering and Borden's (1962) study showed that the tidal wind vector usually turns clockwise at an approximately constant rate, and its amplitude does not vary greatly through the course of the day. Our own results presented in the next section bear this out. Hence, the 12-hr wind difference for a single pair of hours should serve to determine fairly accurately both the amplitude and phase of the diurnal tide. A notable exception to this circular turning of the wind vector occurs near the earth's surface, where it has been observed to describe elongated

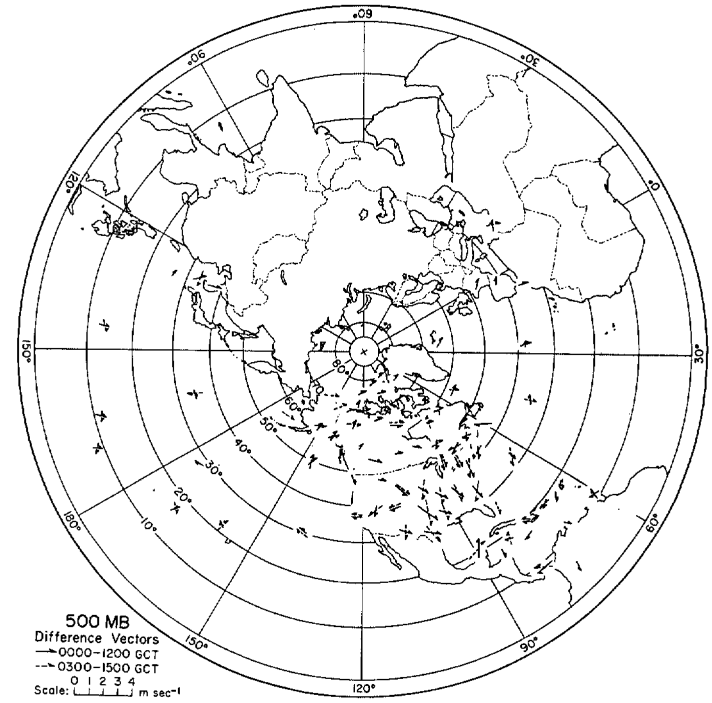

FIGURE 2.—Same as figure 1 for 500 mb.

ellipses in some areas. This should be borne in mind when interpreting the wind difference data at 900 mb (figs. 1A[2] and 1B).

3. HORIZONTAL DEPICTION OF THE TIDAL WIND FIELD

The mean annual wind difference vectors for the Northern Hemisphere stations have been plotted on constant pressure charts (figs. 1A–6) with the center of the wind vector over the station.[3] In the troposphere, two vectors were plotted for most stations. The solid vector indicates the 0000–1200 GMT wind difference and the dashed vector the 0300–1500 GMT wind difference. Due to the low density of data in the stratosphere during the years prior to 1957, the 0300–1500 GMT difference vectors have not been plotted above 200 mb. Length of the difference vector is proportional to the amplitude of the diurnal wind difference with appropriate scales given on the figures. Figure 7 presents similar data at 500 mb and 100 mb for the Antarctic Continent.

It should be emphasized that the maps show an instantaneous picture of the tidal oscillation over the

[2] The 900-mb diurnal wind differences over the Pacific Ocean shown in figure 1A are typical of the layer between the surface and 700 mb. There is a marked divergence associated with this pattern, particularly in the latitude band between 10° N. and 25° N. in the western Pacific. In this region, the diurnal cycle in divergence has an amplitude on the order of 5×10^{-7} sec^{-1} with divergence occurring during the day, convergence at night. This amplitude is within an order of magnitude of the size of the divergence fluctuations believed to accompany the passage of easterly waves in the Tropics. Therefore, it seems quite possible that these diurnal wind fluctuations are an important factor in explaining the nighttime maximum in precipitation observed over large regions of the tropical oceans (Kraus, 1963; Finkelstein, 1964; Lavoie, 1963). For further evidence of the relation between low-level tidal convergence and diurnal variations in cloudiness and precipitation, the reader is referred to a recent study by Brier and Simpson (1969).

[3] Maps for many of the intermediate levels are available upon request from the authors.

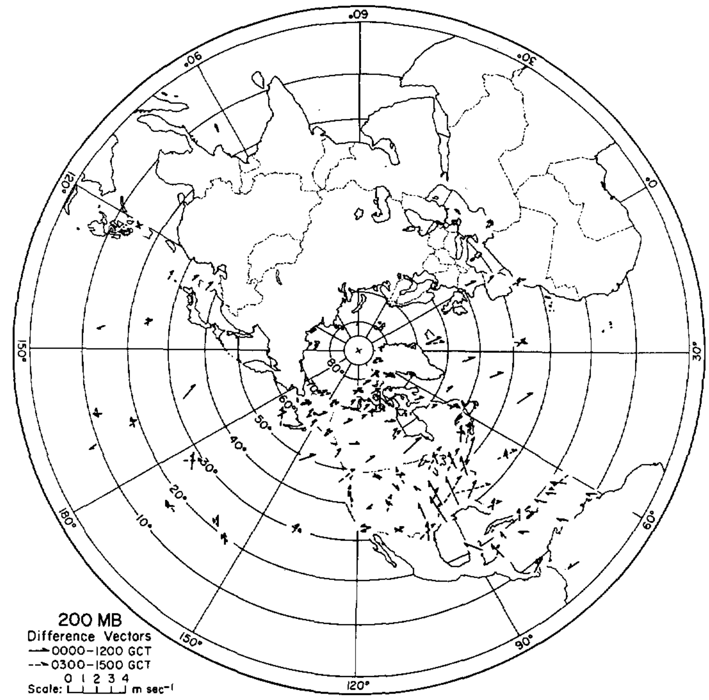

FIGURE 3.—Same as figure 1 for 200 mb.

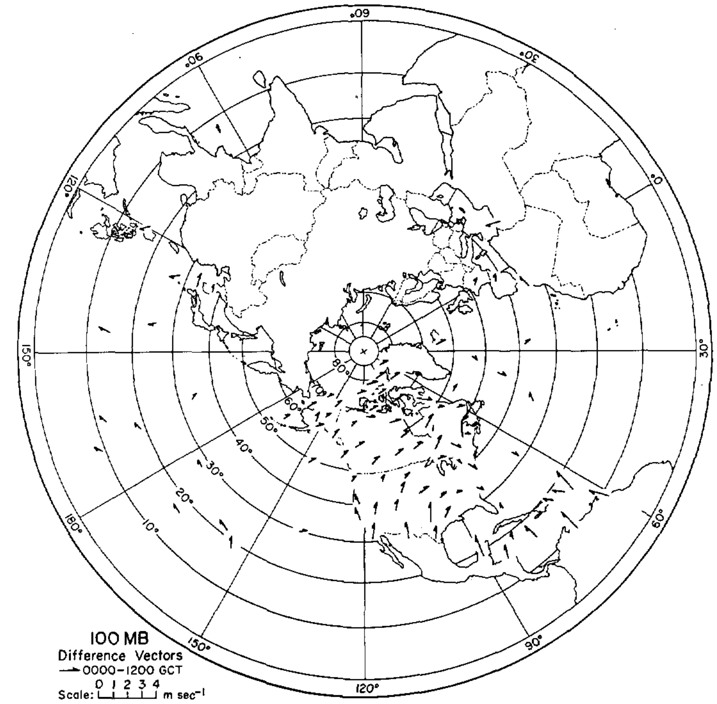

FIGURE 4.—Annual average 0000–1200 GMT wind differences at 100 mb plotted in vector form.

complete hemisphere, and thus local time will vary with longitude. For example, a 0000–1200 GMT wind difference will be the midnight minus noon observation over Western Europe, noon minus midnight along 180° longitude, and sunset minus sunrise over the central United States.

Various broad-scale features of the diurnal tide become apparent through horizontal depiction. Among these are the following:

1) The phase of the tide in the polar regions is independent of height, with the oscillation crossing the Poles from local noon toward local midnight in figures 2–6. The instantaneous longitudinal dependence of the wind vector is given by wave number 1. Figure 7 shows a similar pattern over high latitudes of the Southern Hemisphere. (The same behavior is observed at all levels above 700 mb.) The fact that this agrees with the results of Finger, Harris, and Teweles (1965) probably reflects the dominant influence of the higher latitude stations in their analysis

2) The tidal regime in the midlatitude and tropical regions is more complicated than in the polar regions. Amplitude and phase both vary with height. The instantaneous wind vector no longer exhibits a simple uniform clockwise turning with longitude as is observed in the polar regions, but appears to be controlled largely by topographical influences. Largest amplitudes tend to occur over land in subtropical latitudes.

3) In the lowest levels depicted, a complex behavior is evident. A broad-scale but small-amplitude oscillation prevails over the Pacific Ocean (fig. 1A) where the wind vector turns through 90° in 60° of longitude reflecting a space scale smaller than wave number 1. A much more intense regime is found over North America (fig. 1B). The spatial variation of the wind vectors appears to be related to the major topographical features of the continent.

4) High-altitude patterns, on the other hand, show the oscillation to have comparable amplitude over both land and sea. Patterns are on the scale of continents and oceans. Note the gyre centered over the North American Continent in figure 6.

5) Mountain ranges influence the diurnal tide throughout the troposphere. The Rocky Mountains of North America affect the oscillation as high as the 200-mb level (figs. 1–3), much higher than the elevation of the mountains themselves. Above 200 mb (figs. 4–6), the effects of the mountains disappear, and the oscillation smooths out to continental scale.

6) Large-scale convergence and divergence patterns are apparent. At the 500-mb level, the wind difference vectors over the West Indies appear to converge with those over the southeastern United States at 0000 GMT (fig. 2). This divergent character of the tidal wind field was pointed out by Rasmusson (1967), who stressed its importance in relation to diurnal variations in precipitation.

7) In the Northern Hemisphere, there is a general clockwise turning of the wind difference vector with time. Above the 900-mb level (figs. 2–3), most of the 0000–1200 GMT and 0300–1500 GMT difference vectors intersect each other at angles of about 45°, suggesting that the tidal wind vector turns uniformly with time. As pointed out above, under these conditions, the wind difference vector may be used to estimate the amplitude and phase of the tide.

Wind difference vectors for the 0000–1200 GMT at the 60-mb level were plotted for the summer months, June–

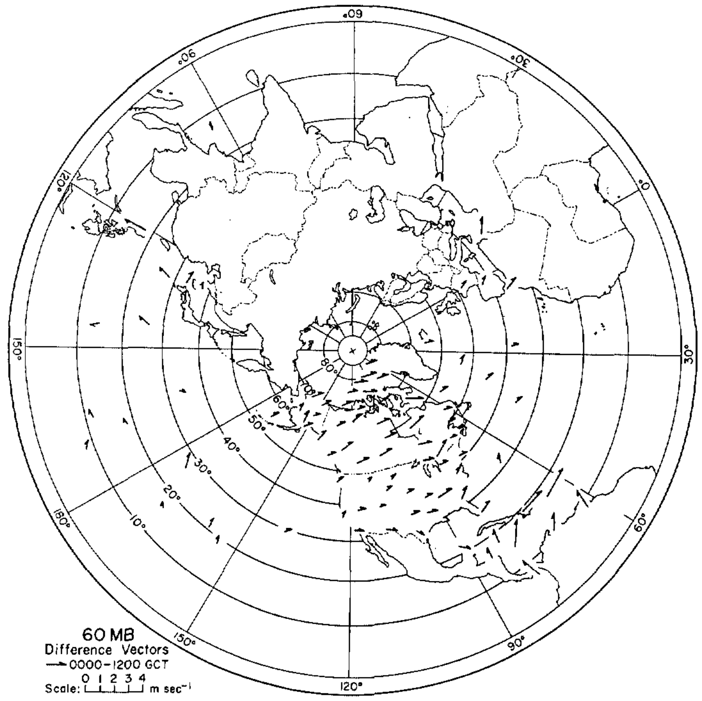

FIGURE 5.—Same as figure 4 for 60 mb.

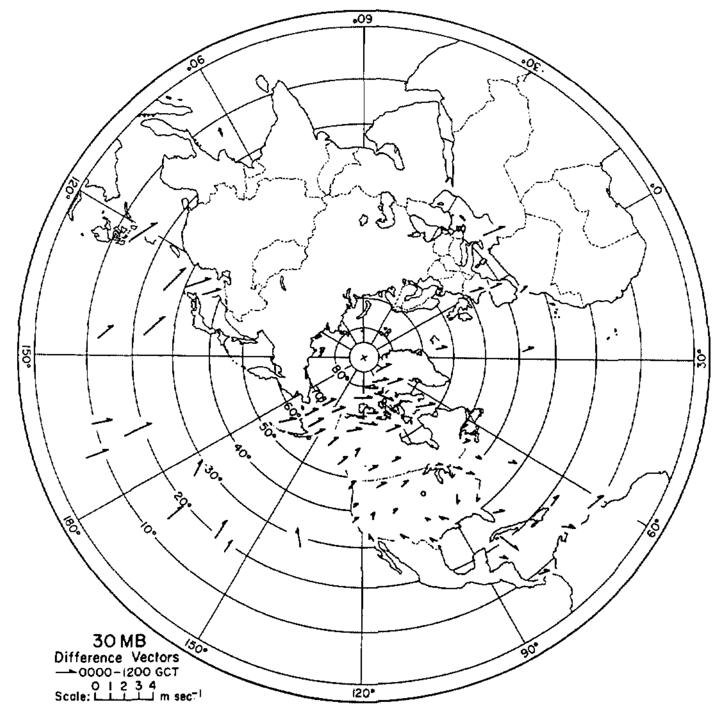

FIGURE 6.—Same as figure 4 for 30 mb.

August, in figure 8 and the remaining 9 mo in figure 9. The following seasonal dependence is apparent from the diagrams.

1) The wind difference vectors over the Poles at both levels indicate no seasonal change of phase. The amplitudes of the wind differences for the summer season are larger, but still of the same order of magnitude as for the rest of the year.

2) Midlatitude areas appear to be highly seasonally dependent. The contribution of the summer season apparently accounts for almost all of the topographically dependent features that appear in the annual means. The same seasonal dependence was noted in analyses at the 500-mb level (not shown).

4. REVIEW OF CURRENT TIDAL THEORY

The solutions of the linearized governing equations are separable with respect to the latitudinal, longitudinal, and vertical coordinates. In treatments that neglect topography, such as Lindzen's (1967), the longitudinal dependence of the diurnal tide is generally assumed to be in the form of a westward propagating wave with zonal wave number 1. The latitudinal dependence is then given by the sum of a series of Hough functions where the coefficients of the respective terms in the series are determined by the latitudinal distribution of the tidal forcing function. Each of the Hough functions is related by its separation constant to a characteristic vertical structure function.

Lindzen (1967) showed that there are two sets of Hough functions that may be excited by an arbitrary distribution of forcing. One set is related to vertically propagating waves, which transport energy away from the level of forcing. In general, the phase propagation is in the opposite direction from the energy propagation. For upward energy propagation this requires that at a particular instant in time the tidal wind vector rotate clockwise with height in the Northern Hemisphere, counterclockwise with height in the Southern Hemisphere. Away from levels of forcing and dissipation, energy density is conserved and, therefore, amplitude increases with height as the square root of the density decreases. The other set of Hough functions is characterized by an exponential height dependence. Phase does not vary with height, and energy does not propagate away from the level of forcing; thus the term "trapped" modes. Figure 10 contrasts the appearance of the two modes, in vertical profile and on a hodograph.

The vertically propagating modes attain their largest amplitudes at low latitudes and have very small amplitudes poleward of about 45°, whereas the trapped modes have largest amplitudes at the higher latitudes.

McKenzie (1968) extended Lindzen's treatment to include the effects of topography. He assumed a westward propagating tidal wave, whose amplitude is a function of longitude. The solutions were still obtainable in the form of Hough functions with the same properties as those described above.

5. COMPARISON OF RESULTS WITH THEORY

HIGH LATITUDES

A quantitative comparison between the observational results and Lindzen's (1967) model was possible only in

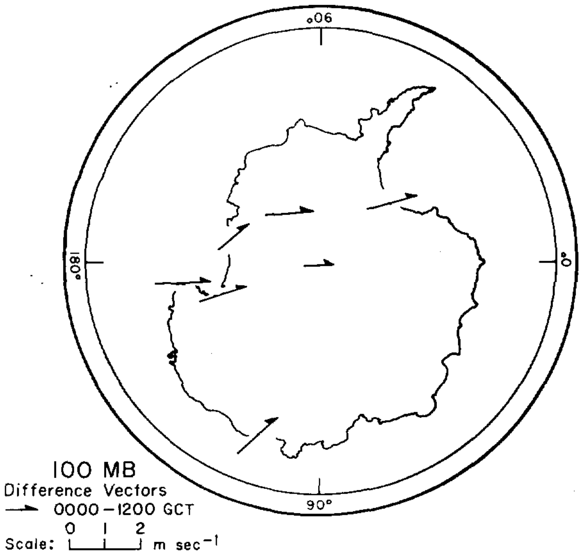

FIGURE 7.—Same as figure 4 for 100 mb.

FIGURE 8.—Average 0000–1200 GMT wind differences for the summer months (June–August) at 60 mb plotted in vector form.

the polar regions where topographical influences are not important. At these latitudes, the phase of the tide is in complete agreement with Lindzen's model, with the maximum poleward component occurring at local noon and the maximum westerly component 6 hr later. The behavior at these latitudes reflects the dominance of the trapped modes, in which phase is independent of height.

To compare the observed amplitudes with Lindzen's, the 0000–1200 GMT difference velocities of all stations north of 60° N. were averaged together at each level. The Antarctic diurnal velocity differences were averaged in a similar manner for levels above 500 mb. (Below 500 mb the phases were irregular.) Theoretical tidal amplitudes based on an isothermal atmosphere were extracted from tables kindly supplied by Lindzen. Figure 11 compares the theoretical and observed 12-hr wind differences for the polar regions. The range of the theoretical tide is one-third to one-half that of the observed at low levels and about 25 percent smaller than the observed at 10 mb. The close agreement between the observed Arctic and Antarctic amplitudes is noteworthy.

The seasonal variations in the tide at high latitudes (figs. 8–9) agree fairly well with Lindzen's results, to the extent that they can be compared. Phase does not vary with season, and summer amplitudes are generally somewhat larger than those for the rest of the year. A quantitative comparison of observed and theoretical amplitudes for the summer and winter seasons cannot be made because Lindzen's values are given for the solstices and equinoxes, whereas ours are averaged for the periods June–August and September–May.

LOW AND MIDDLE LATITUDES

At lower and middle latitudes, topography is clearly the dominant influence on the tidal wind field. Before attempting to compare the topographically forced tide with McKenzie's numerical results, it will be helpful to examine the vertical structure of the tide at these stations. The diurnal tide at a given location reflects the superposition of disturbances generated by local, regional and continental-scale features as well as the portion of the tide that is independent of topography. Hence, the vertical profile of the diurnal wind differences observed over an individual station may be expected to be quite complex. Nevertheless, there are certain generalizations that can be made from examining the vertical profiles for all available stations:

1) At levels above 200 mb, the hodographs for all stations within 45° of the Equator show evidence of only vertically propagating modes. These modes are characterized by a clockwise turning with height in the Northern Hemisphere, counterclockwise in the Southern Hemisphere. This is indicative of upward energy propagation and hence of an energy source within the troposphere. Figures 12–13 show sample hodographs from a selection of typical stations.[4]

2) The tidal fluctuations at low levels can be explained on the basis of trapped modes, which decay with increasing height (fig. 14), or a superposition of trapped and propagating modes (fig. 15). The propagating modes are particularly evident in the Great Plains Region of the United States (fig. 15). With the exception of El Paso, San Antonio, and Balboa, all hodographs at low and middle latitudes show a clockwise turning with height in the Northern Hemisphere (counterclockwise in the

[4] A wider selection of hodographs is available upon request from the authors.

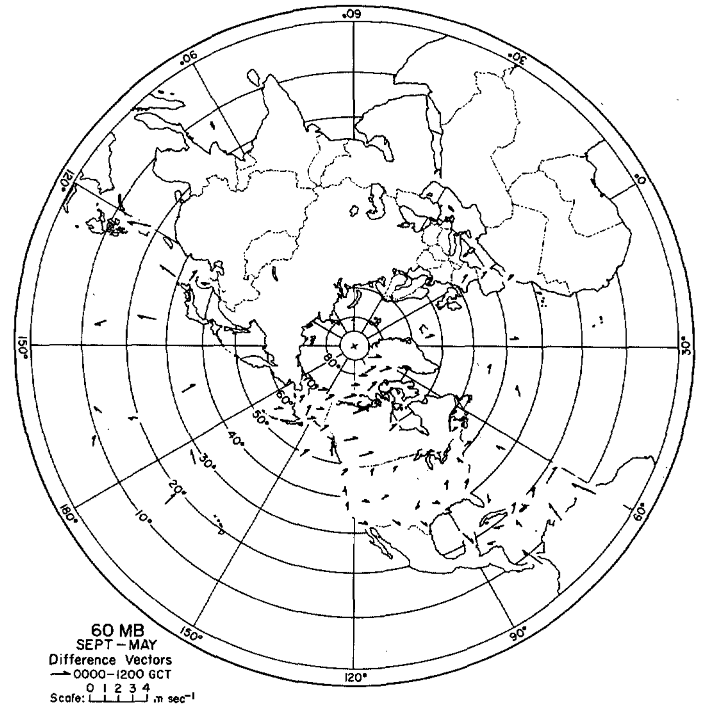

FIGURE 9.—Average 0000–1200 GMT wind differences for the months September–May at 60 mb plotted in vector form.

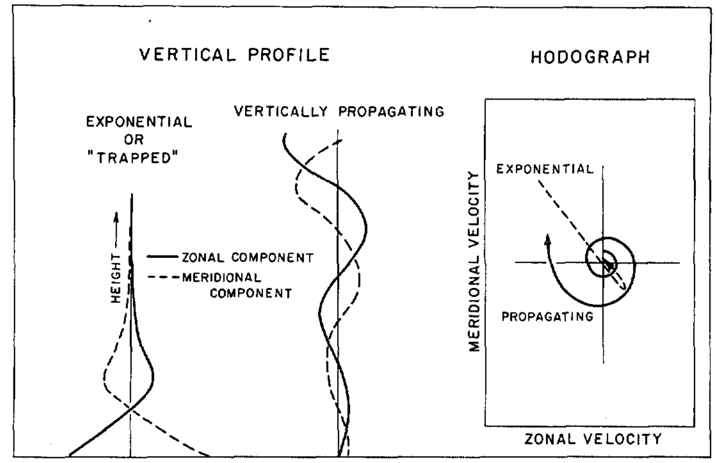

FIGURE 10.—Idealized vertical profiles and hodographs of 12-hr wind differences, illustrating downward propagating (Northern Hemisphere) and trapped modes.

Southern Hemisphere) above the lowest kilometer. This is indicative of upward energy propagation.

3) The weather ships and the stations on small islands far removed from large landmasses do not exhibit any significant topographically induced tidal features. The only islands large enough to generate appreciable tidal effects of their own are the Hawaiian Islands, Cuba, Puerto Rico, Jamaica, and the Philippines.

4) Stations at latitudes above 45° do not exhibit large topographically related features.

5) The propagating modes excited by an isolated topographical feature tend to disperse with increasing altitude. The most convincing evidence of this is the fact that the tidal wind patterns in figures 1–6 become simpler with increasing altitude, with only continent-ocean-scale features reaching the stratosphere.

The strong predominance of downward phase propagation in the vertically propagating modes suggests that the energy source of the topographically generated tide lies in the planetary boundary layer. This is not surprising, in view of the large diurnal temperature oscillation which takes place over land at these levels.

By assuming a low-level heating oscillation based on an idealized continent-ocean distribution, McKenzie (1968) was able to simulate in his model many of the features in the observational data. The shapes typically found on the hodographs were represented well, and the theoretical amplitudes were of the proper order of magnitude and showed a tendency to decrease with latitude poleward of 30°, as is observed. McKenzie's solutions continued to increase in amplitude with height above 30 km, reaching amplitudes in excess of 10 m sec^{-1} above 40 km. This suggests that topographically generated tides might account for a substantial fraction of the observed tide at rocket and meteor wind levels. This is contrary to indications based on the limited amount of rocket data thus far analyzed (Reed, 1968).

COMMENTS ON DARKOW AND THOMPSON'S MODEL

The theoretical treatments described above assumed a stably stratified atmosphere in hydrostatic equilibrium. Darkow and Thompson (1968) attempted to model the diurnal wind oscillation over the central United States by imposing a vertical motion fluctuation at the lower boundary of a homogeneous nonhydrostatic atmosphere, subject to the assumption of vanishing vertical velocity at the tropopause. This arrangement resulted in standing wave solutions in the vertical, which resemble, in some respects, the observed tide at some of the stations in that region. Apart from the question of the applicability of these results to the real atmosphere, which is stably stratified, there appear to be some important inconsistencies between Darkow and Thompson's conception of the tide and the observational results presented above.

1) The diurnal tide is definitely present at stratospheric levels over the central United States. There is no basis for the *a priori* assumption that vertical motion vanishes at the tropopause or any other level. The vertical discontinuities in static stability observed in the atmosphere are not large enough to cause any appreciable reflection of wave energy and therefore standing waves do not appear to be an appropriate description of the vertical structure of the tide.

2) Darkow and Thompson's mathematical formulation specifies an abrupt change in the character of vertical profiles from standing waves poleward of 30° latitude to exponential solutions equatorward of that latitude. The

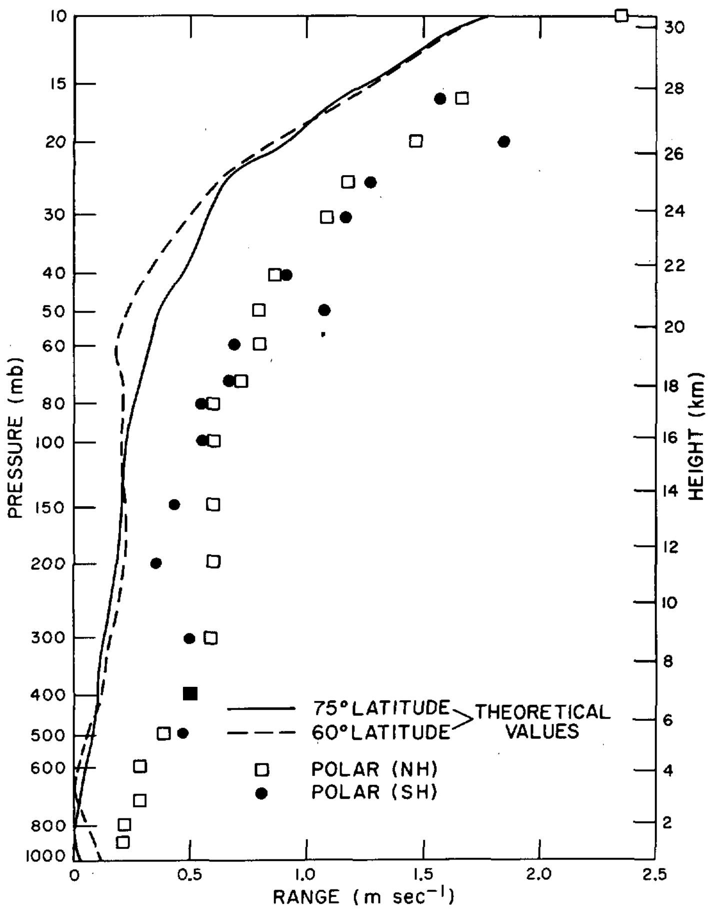

FIGURE 11.—Comparison of theoretical and observed values of tidal range (2 × amplitude) in polar regions. The data points were obtained by averaging data for all stations poleward of 60°. Theoretical values are from Lindzen's (1967) tide tables.

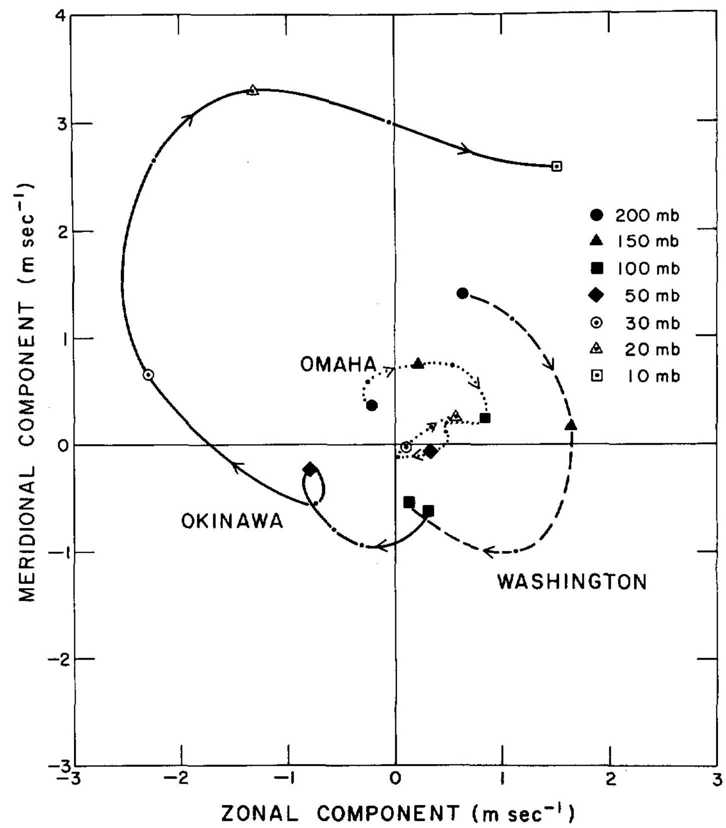

FIGURE 12.—Hodographs of 0000–1200 GMT wind differences above 200 mb for selected Northern Hemisphere stations. Wind differences are directed from the origin to the data points. Data for Okinawa are omitted below 100 mb to avoid congestion near the origin. All hodographs curve toward the right with increasing height, illustrating the dominance of the downward-propagating modes. (Compare with fig. 10.)

observations show quite the opposite—vertically propagating waves tend to be concentrated at lower latitudes while an exponential behavior is observed near the Poles. There is no abrupt discontinuity at 30° latitude.

6. TOPOGRAPHICAL INFLUENCES UPON THE DIURNAL TIDE

Several mechanisms have been suggested whereby topography may influence the character of tidal oscillations. Perhaps the most familiar example is the sea breeze, in which horizontal variations in the heat storage capacity of the earth's surface give rise to horizontal gradients in the amplitude of the diurnal temperature cycle in the planetary boundary layer. The effect of land-sea differences upon the 12-hr wind difference field can be seen by noting that most of the wind difference vectors at 900 mb (fig. 1B) at stations on the coastal plain, with the exception of those in the south-central United States, are consistent with onshore flow during the afternoon. The Atlantic coastal sea-breeze pattern appears to extend well inland over the northeastern United States. Even at stratospheric levels (e.g., 30 mb, fig. 6), the wind difference patterns reflect the continent-ocean influence.

A uniform heating cycle on sloping terrain also produces oscillating horizontal temperature gradients. Lettau (1967) and Holton (1967) have suggested that this may be the cause of the large oscillations in the low-level wind fields over the west slopes of the Andes and the Central Plains of the United States. The wind difference vectors at 900 mb over both regions are indicative of upslope motion in the late afternoon, which is consistent with this type of forcing. The failure of the sea breeze on the Texas coast to penetrate inland to Shreveport and San Antonio (fig. 1B) may be due to the dominance of the effect of terrain slope in this region, which causes upslope flow toward the highlands of central Mexico during the afternoon hours.

Quasi-stationary boundaries between air masses, edges of persistent cloud sheets, and changes in the type of vegetation from one region to another are other factors that may contribute to the observed diurnal oscillation since they produce horizontal contrasts in the amplitude of the diurnal temperature cycle.

Buajitti and Blackadar (1957) proposed a quite different type of mechanism to explain the low-level wind fluctuations in the Great Plains. They suggested that the change in eddy viscosity in the boundary layer between day and

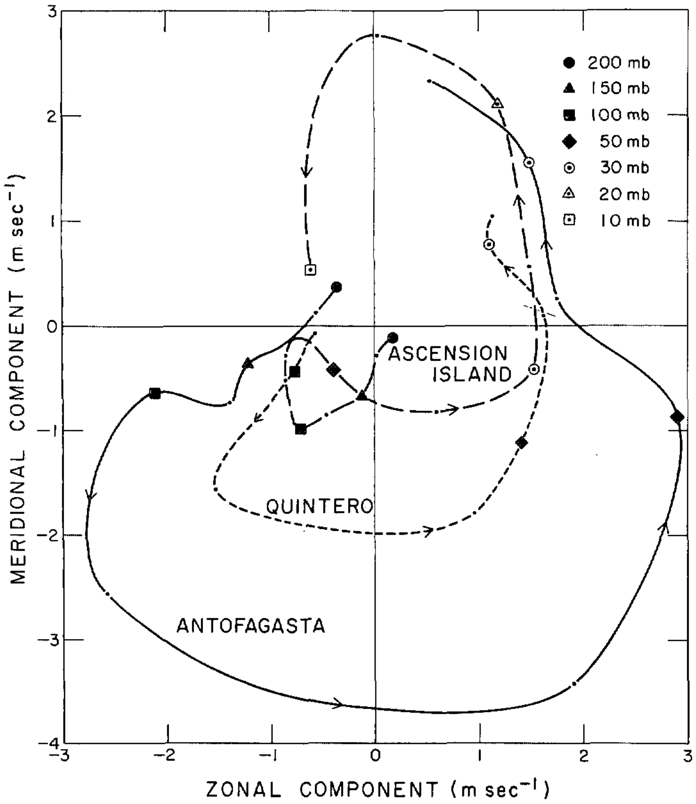

FIGURE 13.—Hodographs of 0000–1200 GMT wind differences above 200 mb for selected Southern Hemisphere stations. Wind differences are directed from the origin to the data points. Data for Quintero below 125 mb are omitted to avoid congestion near the origin. All hodographs curve toward the left with increasing height, illustrating the dominance of the downward-propagating modes. (Compare with fig. 10.)

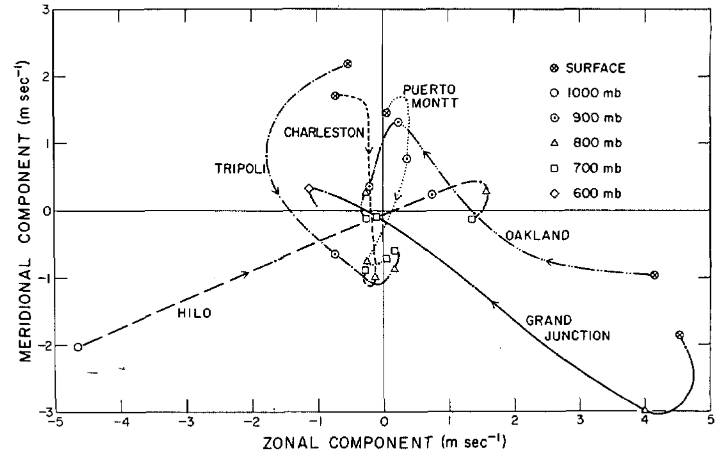

FIGURE 14.—Hodographs of 0000–1200 GMT wind differences at the lowest six reporting levels for selected stations which show exponential (trapped) behavior at low levels. (Compare with fig. 10.)

night over landmasses gives rise to a time-varying frictional force, which interacts with the steady-state (topographically influenced) low-level wind field to produce an inertial oscillation. There is little doubt that frictional forces exert a modifying influence upon the 12-hr wind difference field within the atmosphere's lowest kilometer where many of the hodographs in figures 14–15 exhibit marked departures from the behavior that would be expected if heating alone were important.

On the basis of the data presented in this paper, it is not possible to determine whether friction is responsible for the strong influence of topography upon the tide, but in view of the strong correspondence between the tidal wind field and heating patterns, it appears that it may be of only secondary importance in many regions.

7. INTERACTION BETWEEN THE TIDE AND SYNOPTIC DISTURBANCES

Regardless of whether the diurnal tide in the troposphere is directly driven by solar heating, or whether it is a result of the change in eddy viscosity which heating produces, it is clear that the response of the atmosphere to the diurnal heating cycle must be influenced by synoptic disturbances. The day-to-day changes which these systems produce affect the distribution of cloudiness, humidity, static stability, and surface wind, any or all of which may alter the tidal wind patterns. When the tide-generating mechanisms are sufficiently well understood, it should become possible to predict day-to-day amplitudes of the tide with greater accuracy than the pure climatological approach would afford. Such a development might significantly improve short-range forecasting in the Tropics and over middle latitudes during summer.

It is also quite possible that the reverse type of interaction takes place, i.e., that the tidal fluctuations exert an influence on synoptic patterns. Kung (1966) showed that the conversions between available potential and kinetic energy at certain levels vary significantly between 0000 and 1200 GMT in the annual average over North America. The widespread poleward (down gradient) motion in the 0000–1200 GMT wind differences at 900 and 200 mb and the reverse flow at 500 mb coincide with the levels where Kung obtained his largest diurnal differences in energy conversion. The cross-gradient tidal flows shown in figures 1B, 2, and 3 are in the proper direction and of the proper order of magnitude to account for Kung's results. It is not possible to say to what extent this "sloshing back and forth" between available potential and kinetic energy is truly reversible. In view of the high degree of nonlinearity inherent in atmospheric motions, it is possible that inclusion of the tide in numerical prediction models may have a significant effect on forecasts beyond 1 or 2 days.

Aside from its possible application to forecasting, further study of the tide may provide valuable insight into the nature of diabatic heating and frictional dissipation within the planetary boundary layer and their effect upon the general circulation.

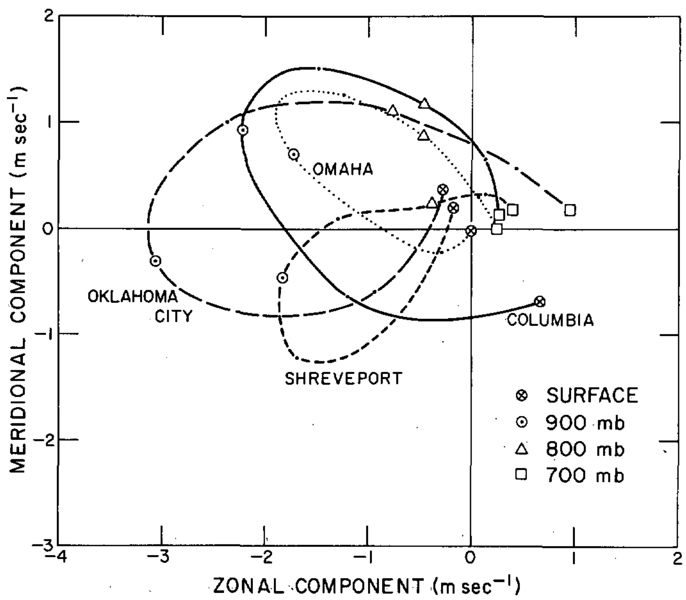

FIGURE 15.—Hodographs of 0000–1200 GMT wind differences at the lowest six reporting levels for stations which show mixed (downward propagating and trapped) behavior at the low levels.

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